Hydraulic fracturing during the formation and deformation of a basin: A factor in the dewatering of low-permeability sediments

John W. Cosgrove

ABSTRACT

The geological expression of hydraulic fracturing is varied and is controlled primarily by the magnitude of the differential stress and the intrinsic properties of the rock. The orientation and type of fractures that develop within a basin are determined by the state of stress, which in turn is controlled by the geological boundary conditions. During the early stages of burial and diagenesis the formation of hydraulic fractures is thought to be an important factor in the movement of fluids through and out of low-permeability, semilithified sediments. Unfortunately, these fractures are not generally preserved and are presumed to heal once the fluid pressure is relieved.

The low-permeability Mercia Mudstones of the Bristol Channel Basin, southwest England, however, contain bodies of sand that, during the opening of the basin, were injected along some of the hydraulic fractures in the mudstones, preserving them as sedimentary dikes and sills. Field observations indicate that fluid pressures within the Mercia Mudstones were also very high during basin inversion and that hydraulic fracturing provided a transient permeability that relieved this excess pressure. The fractures are not visible in most of the mudstones but have been preserved within evaporite-rich horizons as a network of satin spar veins. Thus, the chance preservation of the sedimentary dikes and satin spar veins shows that at different times during the evolution of the basin, fluids migrated through low-permeability units along transient networks of hydraulic fractures. In addition, the orientation and spatial organization of these fractures reflect the boundary conditions operating at various stages in the basin history.

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INTRODUCTION

The movement of fluids through low-permeability sediments and rocks and the associated processes of fluidization and brittle failure by hydraulic fracturing concern a large group of geoscientists ranging from hydrocarbon and mining geologists attempting to understand the passage of hydrocarbons and mineralizing fluids through low-permeability rocks over millions of years to engineering geologists interested in the short-term response of water-saturated sediments and rocks to seismic events. In this article I describe geological evidence for the formation of transient hydraulic fracturing. This evidence takes the form of sedimentary dikes and satin spar veins and comes from the Bristol Channel Basin situated off the north Somerset coast in southwest England. Their preservation is a direct result of the type of sediments within the basin, and these reflect the arid environment prevailing during the infilling of the basin.

I illustrate how, by using simple mechanical principles, it is possible to predict the type and orientation of fractures that might develop within a dewatering sedimentary succession while it is situated in the various boundary conditions associated with the different stages of basin evolution.

This article begins with a brief discussion of brittle failure, hydraulic fracturing, and the stress states thought to exist in the crust during basin evolution. The orientation and spatial distribution of hydraulic fractures that might be expected to form at different times in the basin history are then modeled, and field evidence is presented to support the model.

BRITTLE FAILURE

The theories of brittle failure and hydraulic fracturing are discussed in most structural texts (e.g., Price, 1966; Phillips, 1972; Price and Cosgrove, 1990; Engelder, 1993), and only a brief summary of these concepts relevant to the ideas discussed in this article is presented here.

Figure 1a is a summary diagram showing the failure envelope for brittle failure. This is constructed in part by the Navier-Coulomb criteria for shear failure ($\tau = C + \mu\sigma_n$, where τ is the shear stress, C the cohesive strength, μ the coefficient of internal friction, and σ_n the normal stress) and in part by the Griffith criteria of extensional failure ($\tau^2 + 4T\sigma_n - 4T = 0$, where T is the tensile strength of the rock). The diagrams in Figure 1b and c show extensional fractures and shear fractures, respectively, and the orientation of the principal stresses that are postulated to have generated them. These stress states are represented as Mohr circles on Figure 1a, where it is clear that in order for shear failure to occur the Mohr circle must be sufficiently large to touch the shear failure envelope. It follows from the geometry of the failure envelope that this can only occur if the diameter of the Mohr circle (i.e., the differential stress $\sigma_1 - \sigma_3$) is greater than four times the tensile strength of the rock (T). It also follows from the geometry of the failure envelope that the angle 2θ between the normal stress axis and the line joining the center of the Mohr circle to the point A where it touches the failure envelope is the angle between the two conjugate shear fractures (Figure 1c). For extensional failure to occur the Mohr circle must touch the

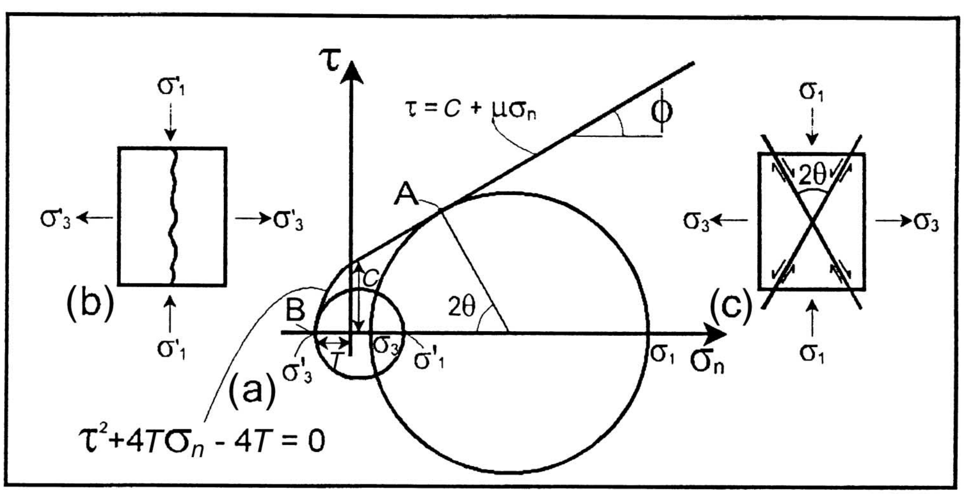


Figure 1. (a) The Navier-Coulomb/Griffith brittle failure envelope. The two Mohr circles represent stress states that give rise to extensional failure (the smaller circle) and shear failure (the larger circle). Parts (b) and (c) show the relationship between the principal stresses and extensional failure and shear failure planes, respectively.

failure envelope at point B. This can only occur if the differential stress is less than four times the tensile strength of the rock.

Thus the type of brittle failure indicates whether the differential stress during fracturing was greater or less than $4T$.

HYDRAULIC FRACTURING

The state of stress in the Earth's crust tends to be compressional, and true tensile stresses are thought to be uncommon. This is particularly true for the stress states in a sedimentary pile undergoing burial and diagenesis in a tectonically relaxed basin. Nevertheless, extensional fractures occur commonly, and this apparent contradiction has been satisfactorily explained by arguing that failure occurs by hydraulic fracturing (see, e.g., Phillips, 1972). It is argued that the internal fluid pressure, p, in the sediment or rock acts so as to oppose the applied stresses and that the rocks respond to the effective stresses $(\sigma_1 - p)$, $(\sigma_2 - p)$, $(\sigma_3 - p)$. Thus a state of lithostatic stress in a rock is modified by the fluid pressure to an effective stress state $(\sigma_1 - p)$ and $(\sigma_3 - p)$, and the Mohr stress circle (Figure 1a) is moved to the left by an amount equal to the fluid pressure. This may bring the circle into contact with the failure envelope, resulting in hydraulic fracturing. If the circle has a diameter greater than $4T$, then it will impact the shear failure envelope and result in shear failure; if it is less than $4T$, it will impact the extensional failure envelope and result in extensional failure.

Various processes have been proposed to account for the buildup of fluid pressure within a sedimentary succession. These have been summarized by Osborne and Swarbrick (1997), who divide them into three categories: (1) increase in compressive stress (i.e., reduction in pore volume) caused by tectonism or disequilibrium compaction (where a fluid cannot be expelled fast enough, for example during rapid burial of a low-permeability material, the pressure of the pore fluid rises above hydrostatic values; this process is known as disequilibrium compaction); (2) fluid volume change during temperature increase (aquathermal pressuring), diagenesis, hydrocarbon generation, and cracking to gas; and (3) fluid movement and processes related to density differences between liquids and gases caused by hydraulic (potentiometric) head, osmosis, and buoyancy. The processes proposed for the increase in fluid pressure within the Mercia Mudstones that led to the formation of hydraulic fractures during the formation and inversion of the Bristol Channel Basin are those of disequilibrium compaction and tectonism.

Having briefly considered brittle failure, the relationship between stress and fractures, and the process of hydraulic fracturing, we can proceed to consider the factors that affect the state of stress in a sedimentary succession and therefore the type and orientation of the hydraulic fractures that might develop in it.

STRESS STATE WITHIN A BASIN

In this section, the state of stress in sediments being buried in a basin is briefly considered to predict the type and orientation of hydraulic fractures that might form. The stress state depends on the material properties of the sediment or rock and on the boundary conditions.

Consider the relatively simple boundary conditions, which affect sediments in a tectonically relaxed basin, that is, one in which the main source of stress is due to the overburden. If the boundary conditions are such that horizontal strains are prevented by the constraints of the rock mass surrounding the area of interest, then it can be shown (see, e.g., Price, 1966) that the vertical and horizontal stresses are related as follows:

$$\sigma_H = \sigma_V/(m - 1) \tag{1}$$

where m is Poison's number, the reciprocal of Poison's ratio. The vertical stress is σ_1, and its magnitude given by

$$\sigma_v = \sigma_1 = z\rho g \tag{2}$$

where z is the depth, ρ the average density of the overlying rocks, and g the acceleration due to gravity. These equations show that if the overburden has a constant density and Poison's number does not change with depth, then the vertical and horizontal stresses increase linearly with depth (Figure 2).

Figure 2 shows that the differential stress $(\sigma_V - \sigma_H)$, also increases linearly with depth. From the discussion of brittle failure given previously it follows that the hydraulic fractures that form in the upper section of the sedimentary pile where the differential stress is less than four times the tensile strength of the rock are vertical extensional fractures opening against the least principal stress σ_3, and the fractures that form at depths where the differential stress is greater than $4T$

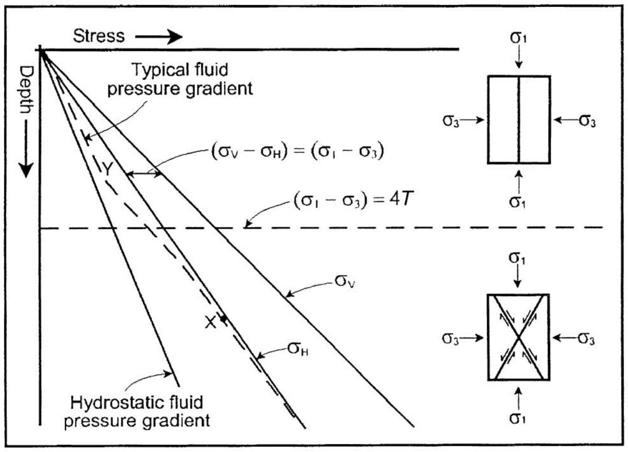


Figure 2. Plot of variation of vertical and horizontal stress and fluid pressure with depths according to equations 1 and 2, which assume that the stresses are generated by the overburden in a tectonically relaxed basin. The differential stress increases with depth. At the depth where it exceeds 4*T*, the fractures change from extensional to shear.

are shear fractures dipping at around 60° (i.e., θ in Figure 1c is 30°). The strike of the hydraulic fractures is controlled by the difference between the two principal horizontal stresses, σ_2 and σ_3. If they are significantly different, the strikes show a marked parallelism with σ_2. If they are the same, the strikes are random (see Cosgrove, 1998).

In the previous discussion I assumed that the state of stress at any depth is determined by equations 1 and 2 and that the density of the rock or sediment and the Poison's number remained unchanged with depth. These assumptions are clearly unreasonable (see, for example, Price [1958] and Eaton [1969] for a discussion of the change of *m* with depth), and direct measurements of the stress state in several present-day basins show that the stresses change in a nonlinear manner with increasing depth (Fertl, 1976; Breckels and van Eekelen, 1982).

Price (1974) suggested that even in a basin experiencing only epeirogenic motion, lateral stresses would be induced in the beds in addition to those caused by the overburden. He argued that these additional stresses depend in part on the geometry of the basin and that if the basin width is greater than a few tens of kilometers, it is necessary to consider the curvature of the earth in calculating these stresses. In addition to the horizontal stresses generated by bed length changes related to the basin geometry, he noted that stresses would be induced by the increase in temperature that accompanies burial.

For a basin that has an elliptical plan undergoing epeirogenic subsidence the equations governing the vertical and horizontal stresses are

$$\sigma_z = z\rho g \tag{3}$$

$$\sigma_y = [\sigma_z/(m - 1)] + Ee_y + E\alpha\Delta t \tag{4}$$

$$\sigma_x = [\sigma_z/(m - 1)] + Ee_x + E\alpha\Delta t \tag{5}$$

where σ_z is the vertical stress, σ_y and e_y are the principal stress and strain along the long axis of the basin, and σ_x and e_x are the principal stress and strain along the short axis of the basin. *E* is Young's modulus, α the coefficient of thermal expansion, and Δt the change in temperature between the sediment-water interface and the depth of interest. Note that equation 3 is identical with equation 1 and equations 4 and 5 similar to equation 2, having extra terms relating to the stress generated as a result of bed-length changes during burial and temperature increase.

These equations show that the magnitude of the vertical and horizontal stresses in a sediment or rock at a particular depth is very sensitive to its material properties, specifically density, Young's modulus, Poison's number, and the coefficient of thermal expansion. Because these properties vary from rock to rock, the state of stress in adjacent beds at any depths can be very different. The stress state in three such beds are shown as Mohr circles in Figure 3a, together with the three failure envelopes. Being at the same depth, all three beds have the same vertical stress. The horizontal stresses are determined by the intrinsic properties of the beds according to equations 4 and 5. Bed (i) has a small differential stress, bed (ii) a larger differential stress but less than four times its tensile strength, and bed (iii) a differential stress that exceeds four times its tensile strength. Thus, if the fluid pressures in the three beds are sufficient to cause hydraulic fracturing, bed (i) will develop extensional fractures that have random dips (i.e., the differential stress is very small, and therefore little constraint on the dip of the fracture exists), bed (ii) will develop vertical extensional fractures (Figure 1b), and bed (iii) will develop conjugate shear fractures dipping at around 60° (Figure 1c).

Whether the fluid pressure necessary for the formation of hydraulic fracturing develops in a bed de-

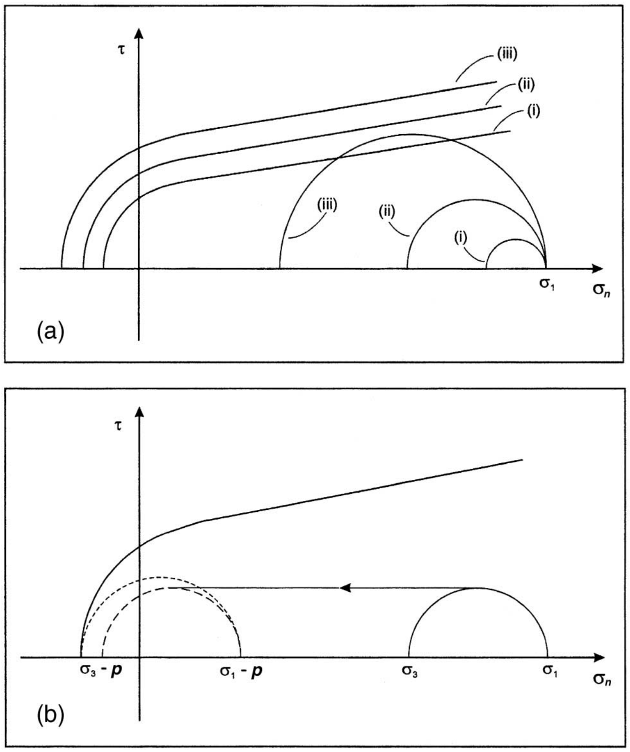


Figure 3. (a) Three stress states in three adjacent rock layers at a depth corresponding to an overburden stress of σ_1. Each rock has its own failure envelope (i), (ii), and (iii). If the fluid pressure becomes sufficiently high to generate hydraulic fracturing in the three beds, bed (i) will develop extensional fractures that have an almost random dip and strike, bed (ii) will develop vertical extensional fractures, and bed (iii) will develop shear fractures dipping at around 60°. (b) Stress states representing the lithostatic (solid circle) and effective stress ($\sigma - p$) (dashed circle) in an overpressured rock in which the fluid pressure is just below σ_3. A small tectonic extension causes a reduction in σ_3 and an increase in the differential stress and is likely to bring the Mohr circle (dotted) into contact with the failure envelope, initiating an episode of hydraulic fracturing and fluid migration.

pends on its porosity and permeability. Thus at any particular depth in such a basin, the fracturing behavior of adjacent beds may be markedly different. Fractures in one bed may be randomly striking and/or dipping extension fractures; in another, vertical extension fractures may form that have either a uniform or random strike depending on the difference between σ_2 and σ_3; in another, conjugate shear fractures that have either a uniform or random strike may occur; and in others there may be no fractures at all. Price (1974) considers the formation of fractures in different rock types during both burial and exhumation, and the interested reader is referred to this original and thought-provoking publication.

The previous discussion makes clear that because of the different properties (Young's modulus, Poison's number, tensile strength, the coefficient of thermal expansion, the porosity, and permeability) of adjacent beds, it is possible to form shear fractures and extension fractures at the same depth. This occurs because the depth at which conditions for extension failure give way to conditions for shear failure (i.e., $(\sigma_1 - \sigma_3) = 4T$) (Figure 1) is different for different rocks.

THE BRISTOL CHANNEL BASIN

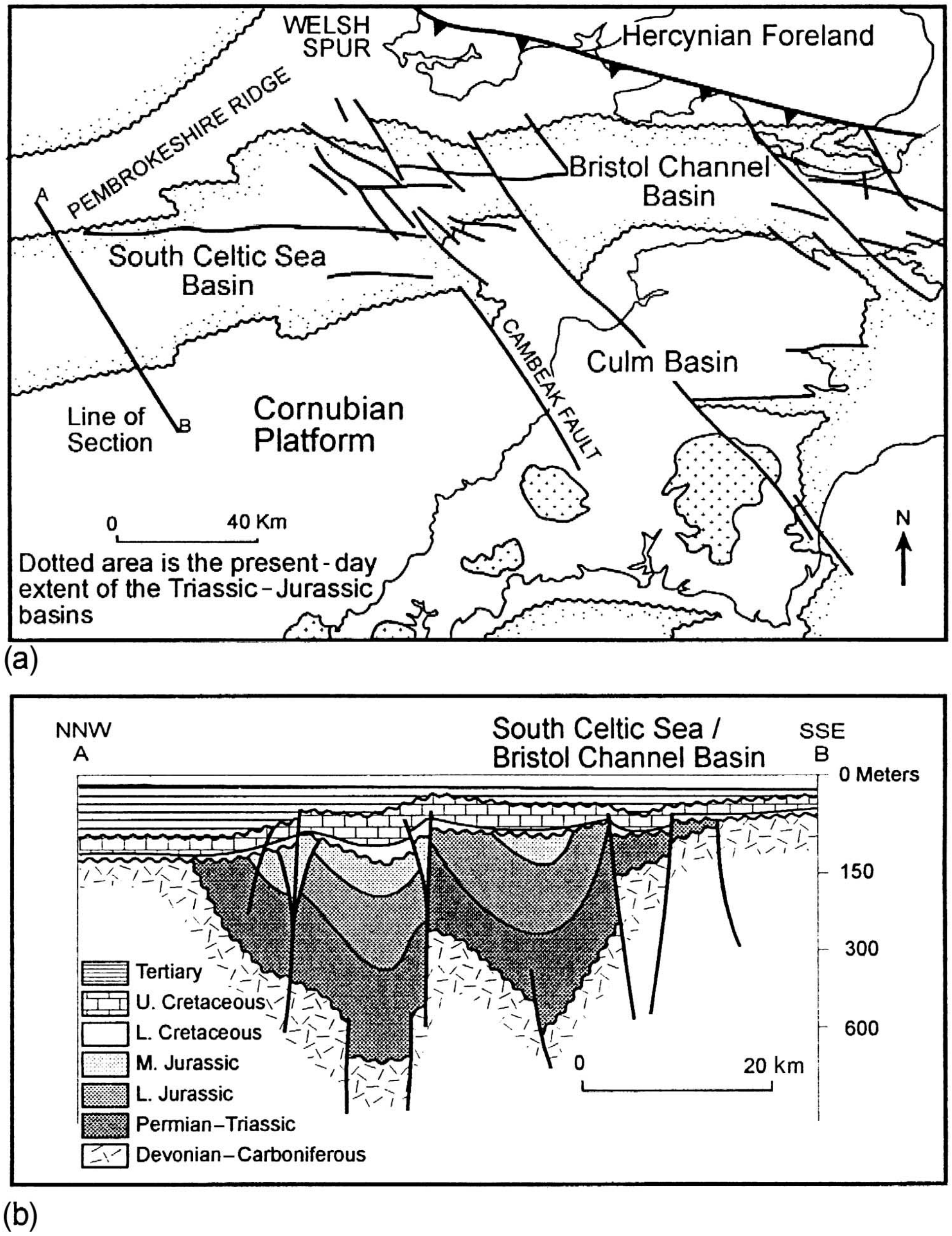
The previous discussion regarding the stress conditions in the crust resulting from an overburden load and the factors that control the orientation and spatial organization of hydraulic fractures can now be applied to a specific basin. The basin selected is the Bristol Channel Basin, and this choice was influenced by the fact that its sedimentary infill is particularly well suited for the formation and preservation of hydraulic fractures.

A detailed discussion of the stratigraphy and tectonic evolution of the Bristol Channel Basin is outside the scope of this article, and the interested reader is referred to Chadwick (1993) and Coward (1995). Only a brief summary is presented here to familiarize the reader with the lithologies in which the hydraulic fractures are formed and preserved and the major stress regimes these sediments experienced during basin opening and inversion.

The Bristol Channel Basin (Figure 4) is part of a framework of intracratonic extensional basins that developed through and around the south of England. These basins were initiated during the Permian and Early Triassic as a result of the breakup of Laurasia. The basins are floored and bounded by Paleozoic rocks of Carboniferous and Devonian age. During the Permian–Triassic the Bristol Channel Basin was situated farther south in the northern desert latitudes, and as a result the sediment infill is typical of an arid environment.

The Permian–Triassic succession commences with continental red beds/alluvial fan deposits. These are coarse-grained conglomerates and breccias, grading laterally and upward into sandstones, which in turn pass into siltstones and mudstones. These are succeeded by coarsely arenaceous Lower Triassic strata of the Sherwood sandstone group, which passes upward into siltstones, mudstones, and a major zone of evaporites of

Figure 4. (a) Map showing the east-west–trending Bristol Channel Basin compartmentalized by northwest-southeast–striking wrench faults. (b) Schematic cross section of the basin (modified from Van Hoorn, 1987).



the Mercia Mudstone Group. A renewal of fault activity in the very Late Triassic and Early Jurassic coincided with a transition from a dominantly continental environment to more open marine conditions. As a result, the sediments of Early Jurassic age form a thick sequence of cyclic shallow-water sediments. These are mudstones and interbedded marine limestones known as the Blue Lias.

Part of the Mercia Mudstones crops out on the north Somerset coast, and in these rocks evidence for the formation of numerous phases of transient hydraulic fracturing can be found. The rock consists of red marl containing several evaporite-rich horizons. These horizons are interpreted as representing periods of basin filling and drying out. Subsequent episodes of rifting result in renewed basin deepening and the deposition of evaporite-free marls. As the basin fills again it dries out, and another evaporite-rich horizon is deposited. Thus, although it might be argued that the evaporite-rich horizons simply represent periods of drying up of the basin, it is also possible to argue that the couplets of evaporite-free marls and evaporite-rich marls correspond to pulses of opening and deepening of the basin followed by infill and drying out. If this interpretation is correct, then the succession of evaporite horizons is a sedimentological record of the minor pulses of extension associated with the opening of the rift system.

In addition to the evaporites, the marls also contain pockets of windblown sand. These pockets vary in size but are typically a few meters thick and a few tens of meters wide. They are composed almost entirely of millet seed quartz grains, and even now these desert-derived sands are only poorly cemented. As is discussed in the following section, the evaporites and the windblown sand are key factors in the preservation of the hydraulic fractures that formed within the marl.

BASIN OPENING

During the opening of the Bristol Channel Basin the regional stress field would have been one associated with rifting. Thus the maximum principal compression would have been vertical, the intermediate principal compression parallel with the basin, and the minimum principal compression normal to the basin edges. During the burial of the marl the overburden stress would increase, and because of its low permeability, the fluid pressure in the marls would increase. The stress profile in the basin would look similar to that shown in Figure 2, except that there would be an additional tectonic extension that would further reduce the minimum principal stress, σ_3. This would move the line representing σ_3 bodily to the left.

This schematic diagram shows that initially the fluid is expelled from the sediment and escapes to the surface. The pore pressure increases by following the hydrostatic pressure gradient. As subsidence continues, however, the permeability of the sediments declines, and at some point fluid starts to be retained. The depth at which this occurs is the fluid isolation depth (point Y in Figure 2), which marks the onset of disequilibrium compaction. If no fluid escapes below the fluid isolation depth, the pore pressure then rises along a pressure-depth path parallel with the lithostatic gradient. Osborne and Swarbrick (1997) note that although the fluid pressure never exceeds the lithostatic pressure at any depth, the pore pressure can exceed the fracture pressure (i.e., the amount of pore pressure a rock can withstand before its tensile strength is exceeded and hydraulic fracturing occurs) where the fracture gradient (the gradient of the line linking the fracture pressures at different depths) is less than the lithostatic gradient.

I propose such a process for the generation of hydraulic fractures during the burial and compaction of the Mercia Mudstones. The orientation of these fractures would be as shown in Figure 2. Fractures generated in the marl during the early stages of burial and diagenesis would be vertical and would strike parallel with the basin margins. These fractures, which would be predicted to occur on all scales, would allow fluids to move through and out of the marls. This would lower the fluid pressure below that needed for hydraulic fracturing, and the fractures would generally close and seal. As burial and tectonism continued, the fluid pressure would continue to rise until the conditions for hydraulic fracturing were reestablished, when the process would repeat.

The arguments outlined previously indicate that as low-permeability sediments undergo burial and diagenesis in a basin, conditions of stress and fluid pressure are likely to be encountered that lead to the formation of hydraulic fractures. Unfortunately, these fractures are generally not preserved, forming as they do at relatively shallow depths where they are unlikely to be preserved as vein systems and where the rock properties are such that barren fractures tend to heal once the excess fluid pressure has been dissipated.

I have examined several quarries and other outcrops of low-permeability shales thought to have been overpressured during their burial and found remarkably little evidence of such fracturing. In places thin (1–2 mm thick) bedding-parallel veins of fibrous calcite that have the fibers oriented normal to the bedding fractures occur, as, for example, in the Kimmeridge shales at Kimmeridge Bay in Dorset. A “chicken wire” texture has been recorded in cores from shales known to have been previously overpressured (Powley, 1990). These are polygonal arrays of fractures along which very thin veins of calcite have been precipitated.

Fortunately, within the Bristol Channel Basin, some of the vertical hydraulic fractures predicted to occur during basin opening are preserved in the Mercia Mudstones as sedimentary dikes. These dikes are fed by the pockets of sand within the marl, and although they are now intensely folded as a result of subsequent burial and compaction, their present geometry (e.g., Figures 5a, 6a) clearly shows that they were initially injected into vertical fractures in the marl.

The effect of a buildup of fluid pressure in uncemented sediments that have no intrinsic cohesion, such as extremely pure sands within a low-permeability matrix, has been discussed by Cosgrove (1995). A

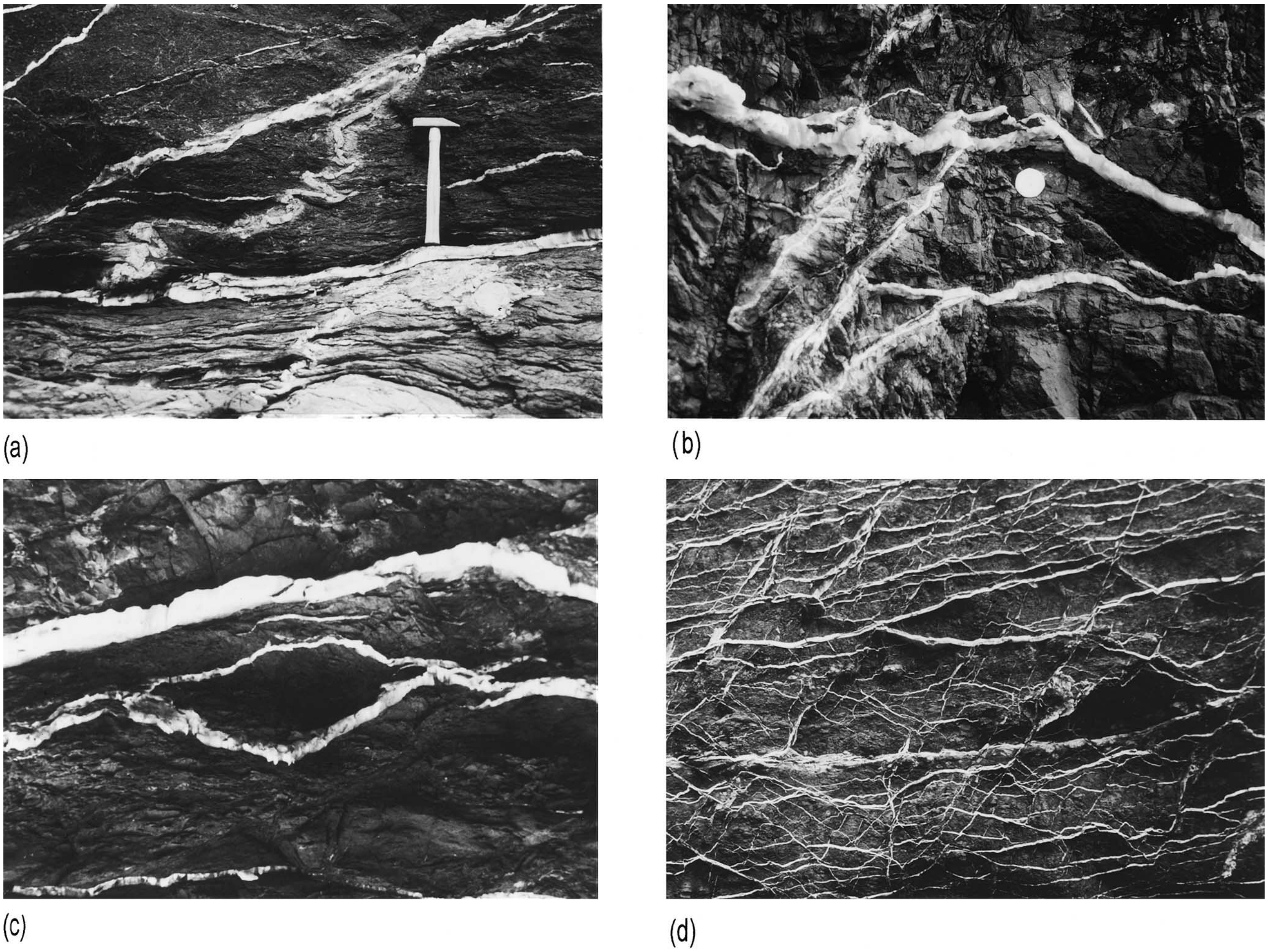


Figure 5. (a) Folded sandstone dike emanating from a pocket of windblown sand and cutting the Mercia Mudstones. (b) Subhorizontal veins of satin spar cutting thin, subvertical sandstone dikes. (c) Satin spar veins formed subparallel with bedding. Note that the fibers of gypsum remain vertical regardless of the local dip of the vein. (d) A pervasive network of satin spar veins recording the pattern of transient hydraulic fractures that developed in the evaporite-rich horizons within the Mercia Mudstones and that, I argue, also developed in the adjacent evaporite-free mudstones where the mineralogy was not suitable for their preservation.

material such as this has no strength and, therefore, cannot support a differential stress. Thus the stress state within it is close to hydrostatic. If the fluid pressure is such that hydraulic fracturing can occur (i.e., if the fluid pressure exceeds the confining pressure), then the grains of the sediment simply move apart slightly and the sediment fluidizes. Because of the electrostatic charges between the clay particles making up the low-permeability matrix, however, an intrinsic cohesion exists in these sediments from the moment they are deposited. This cohesion increases with compaction and cementation. In contrast, the sand lenses within the matrix remain cohesionless until the processes of cementation are initiated. Thus during the early stages of burial and diagenesis the marls possess a cohesion and the sands do not. If during this stage the fluid pressure becomes high enough to cause hydraulic fracturing in both sediments, the marl matrix, because of its cohesion, is able to sustain a differential stress, and, as was discussed previously regarding brittle failure, if the differential stress is less than four times the tensile strength of the marl, extensional failure occurs and the resulting fracture is vertical. At the same time the effect of high fluid pressure on the sand lens is to cause fluidization. Thus regardless of whether the buildup of fluid pressure in the sand lens causes the formation of vertical hydraulic fractures in the marl or the hydraulic fractures generated within the marl propagate and intersect fluidized sand lenses, vertical sandstone dikes form. Subsequent compaction of the marls causes the

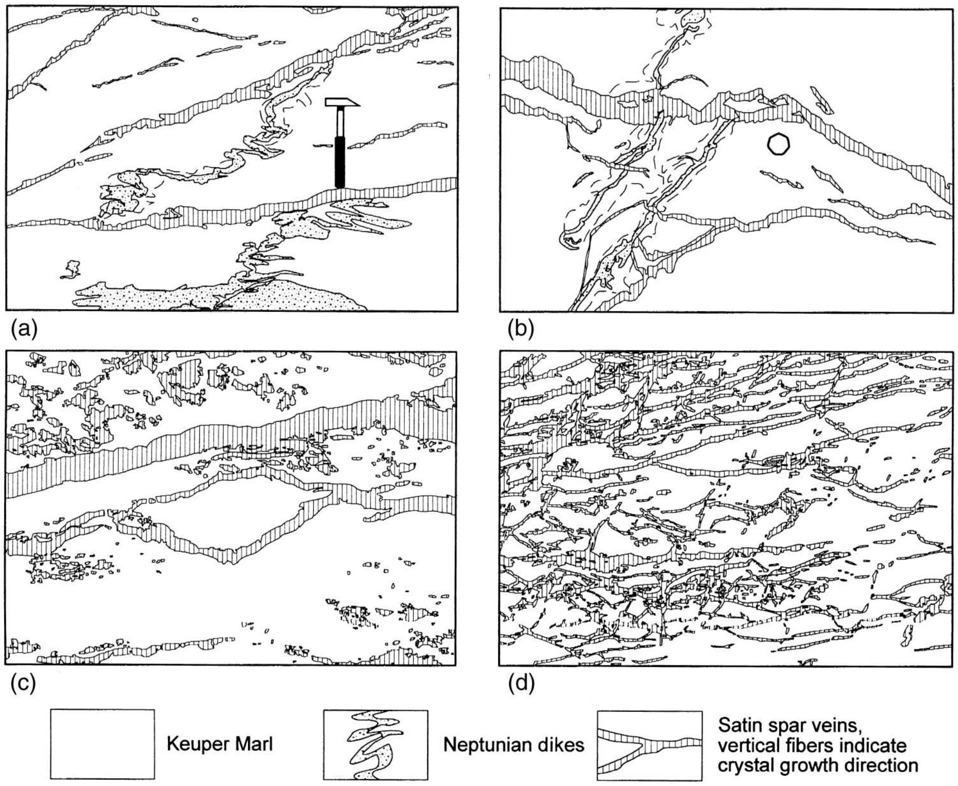


Figure 6. (a–d) Line drawings of Figure 5a–d, respectively.

originally vertical sand dikes to become buckled (Figures 5a, 6a).

Thus, the formation of the dikes preserves the hydraulic fractures and provides proof that the vertical hydraulic fractures predicted from a consideration of the stress state within the sediments during the early stages of burial and diagenesis do occur.

The stress profile shown in Figure 2 and described by equations 1–5 relates to conditions where there is no active tectonic stress. A typical fluid pressure gradient is shown that illustrates the onset of overpressuring and the increase of overpressuring with depth until the fluid pressure approaches the minimum principal stress, σ_3 (point X in Figure 2). The Mohr circles for the lithostatic and effective stress ($\sigma - p$) at this point are illustrated in Figure 3b, which shows that the effective stress is very close to satisfying the conditions of extensional failure. If on further burial the fluid pressure rises sufficiently to overcome the tensile strength of the sediment, then hydraulic fracturing occurs. This enables the fluid pressure to drop below that required for fracturing, and the fractures close and heal. As further burial occurs the fluid pressure increases, and the process repeats. Thus, during burial, a situation arises where overpressured sediments are brought to the point of hydraulic fracturing, after which fluids pass through and out of the sediment in small pulses associated with the episodic formation and healing of hydraulic fractures whose orientation and spatial organization are controlled by the prevailing boundary conditions.

As mentioned previously, the stress profile shown in Figure 2 and represented in Figure 3b is that for a tectonically relaxed basin, that is, one in which the stress field is dominated by the overburden. It is interesting to consider the effect of adding a tectonic stress to this profile on the process of hydraulic fracturing. A relatively simple example is the addition of an episode of regional extension such as may occur during a pulse of opening of a rift basin. The addition of even a small tectonic extension to this stress configuration causes a reduction in σ_3 and a corresponding increase in the differential stress. This change in diameter of the Mohr circle is likely to bring it into contact with the failure envelope (Figure 3b), initiating an episode of hydraulic fracturing and fluid migration. It follows that if a sedimentary succession contained horizons of overpressured sediments that had been brought to the point of hydraulic fracturing and that, as a result of continuing burial, were undergoing periodic hydraulic fracturing and loss of fluid in the manner outlined previously, then even a small episode of tectonic extension could stimulate an important episode of fracturing and fluid migration.

Thus the onset of rifting after a period of thermal subsidence, during which overpressured horizons were brought to the point of hydraulic fracturing and were thus in a state of hydrodynamic equilibrium, is likely to be associated with a major basinwide migration of fluids. Subsidence curves for many rift basins show major pulses of extension followed by periods of tectonic relaxation. A subsidence curve for the Bristol Channel Basin (Figure 7) has been constructed in a generalized form by Beach (1988) to show the estimated thickness of Triassic and Jurassic sequences and the timing and amount of inversion of the basin (see Kamerling, 1979; Beach, 1988). Although Permian subsidence is not shown because no clear estimate of this is available, nevertheless the curves show clearly that the opening of the basin occurred by a series of pulses of extension, each followed by a period of relaxation.

The Mercia Mudstones contain several important evaporite horizons, and I suggested that the couplets of evaporite-free and evaporite-rich marls correspond to pulses of opening and deepening of the basin followed by infilling and drying out. If these couplets do represent a sedimentary record of minor pulses of extension, then these pulses would help maintain the sediments within the basin close to the point of hydraulic fracturing. This would increase the likelihood of major pulses of basin opening (Figure 7), initiating basinwide fracturing and fluid migration.

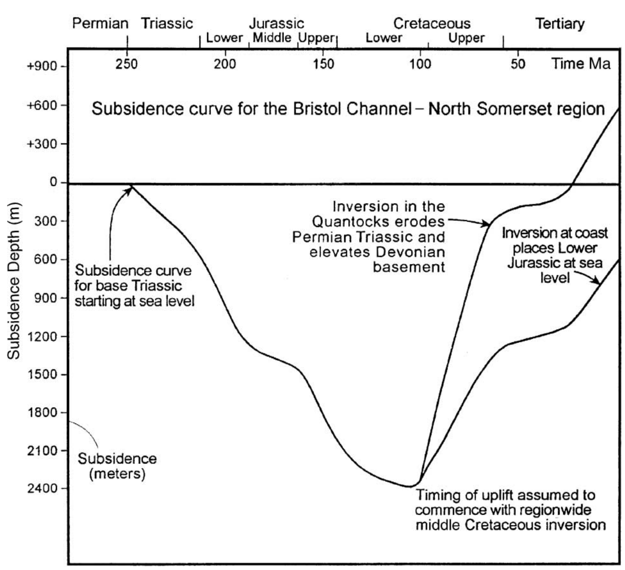


Figure 7. Subsidence curve for the Bristol Channel–north Somerset region, based on stratigraphic data from Kamerling (1979).

BASIN INVERSION

Inversion of the basin began in the middle Cretaceous as a result of a major north-south compression, and the regional stress field experienced by the basin sediments was one in which the maximum principal compression, σ_1, was north-south; the intermediate principal compression, σ_2, was east-west, that is, parallel with the basin margins; and the minimum principal compression, σ_3, was vertical. Thus during inversion, hydraulic fractures would be expected to form either parallel with bedding, if the differential stress was less than four times the tensile strength of the rock and extensional failure occurred, or as shear fractures dipping either 30° north or south, if the differential stress was greater than four times the tensile strength of the rock.

Inspection of the Mercia Mudstones on the north Somerset coast (Figure 4) shows that a complex array of subhorizontal fractures occurs within the evaporite-rich horizons (Figures 5, 6). These fractures are preserved as veins of satin spar, a fibrous form of gypsum, and the orientation of the fibers declares the direction of opening of the veins (see Durney and Ramsay, 1973) The poorly bedded nature of the marl results in the formation of an anastomosing network of veins, which crosscut the folded sand dikes (Figures 5a, b; 6a, b) and therefore postdate them. Figures 5c and 6c show that the fibers remain vertical regardless of the orientation of the vein, indicating that during inversion the fluid pressures were sufficiently high to open the fractures vertically, that is, against the minimum principal stress.

The network of hydraulic fractures is only preserved in the evaporite horizons, and the reason for its preservation relates to the volume changes associated with the hydration of anhydrite to gypsum that occurs as the rock mass is exhumed and the relatively low temperature and pressure conditions necessary for the formation of gypsum are encountered (see, e.g., Jowette et al., 1993). This process was described by Shearman et al. (1972), who pointed out that the hydration of anhydrite to gypsum should result in an increase in volume of 63% if all the calcium sulfate is retained within the system. They note that in many secondary gypsum rocks, former anhydrite is replaced on a volume for volume basis, and the additional volume of gypsum appears as veins that cut the associated rock. Shearman et al. (1972) suggest that in such instances the hydration was caused by water that made entry into the anhydrite-bearing rocks by hydraulic fracture and that the gypsum of the veins grew in the water-filled fractures while the overburden was supported by the water.

For the examples described in the present article, we know that the exhumation during which the hydration occurred was associated with the tectonic inversion of the basin. Thus the basin sediments were experiencing a tectonic compression during their exhumation, which would cause an increase in fluid pressure and facilitate the formation of hydraulic fracturing.

The hydraulic fractures preserved as satin spar veins within the evaporite-rich horizons of the Mercia Mudstones must also have formed in the evaporite-free horizons. Because the appropriate mineralogy for their preservation as veins was not present within the rock, however, they closed and healed as soon as the fluids had past through them and the pressure had dropped. The pervasive network of satin spar veins shown in Figures 5d and 6d provides a glimpse of the transient hydraulic fracture network that must have developed on numerous occasions within the mudstones during exhumation.

DISCUSSION

Considerable debate still exists regarding (1) the properties of sediments during burial and diagenesis, (2) the most appropriate failure criteria to describe them (e.g., Zoback and Healy, 1984), and (3) the conditions necessary to generate fluid pressures of sufficient magnitude to cause hydraulic fracturing (Bowers, 1994; Engelder and Fischer, 1994; Neuzil, 1995; Osborne and Swarbrick, 1997). The aim of this article, however, is not to attempt to resolve these issues but to demonstrate that during the evolution of a sedimentary basin, sediments are brought to the point of hydraulic fracture during both burial and exhumation and that these fractures generate a transient permeability that allows fluids to move through and out of thick, low-permeability successions.

I argue that the fractures generally close and heal once the fluid pressure has been relieved, leaving little or no trace of them having been formed. Fortunately, in the sediments of the Bristol Channel Basin some of these fractures have been preserved as sedimentary dikes and satin spar veins. It is possible to demonstrate that the type (extensional or shear) and orientation of these fractures are compatible with the stress conditions operating at the time of their formation (i.e., the dikes fill vertical fractures associated with basin extension, and the veins fill subhorizontal fractures associated with basin inversion).

The model proposed for the dewatering of the sediments is one in which the sediments are commonly brought to the point of hydraulic fracturing by a combination of disequilibrium compaction and tectonism (Osborne and Swarbrick 1997), the latter associated with episodes of slip on the basin-bounding faults as the basin opens and closes. The effect of a major pulse of tectonism on such a system can be dramatic. For example, the onset of rifting after a period of thermal subsidence during which overpressured horizons achieve a state of hydrodynamic equilibrium, or an increase in the rate of rifting in a rift basin as indicated by the basin subsidence curves (Figure 7), disturbs the state of metastability of the system, during which fluids bleed relatively slowly from the sediments by small episodes of hydraulic fracturing, and is likely to initiate a major basinwide expulsion of fluids.

The distribution and size of the fractures is indicated by the distribution and size of the sandstone dikes and satin spar veins. The dikes (Figures 5a, b; 6a, b) range in width from a few centimeters to a few sand grains. They were originally vertical but are intensely buckled and/or faulted as a result of compaction.

The network of satin spar veins (Figures 5d, 6d), which developed in the evaporite-rich horizons of the Mercia Mudstones, provides a glimpse of the transient fracture network that must have developed on numerous occasions within the mudstones during exhumation. Because the relatively low temperature and pressure conditions necessary for the hydration of anhydrite to gypsum are not encountered until very late in the inversion and exhumation of the basin, the satin spar veins have preserved hydraulic fractures that formed late in the basin's history. Thus the movement of fluids through low-permeability sediments and rocks by the formation of hydraulic fractures is not restricted to burial and diagenesis associated with the early stages of basin evolution, and such fractures clearly can be important factors in the redistribution of fluids within a basin at all stages of its evolution. The vein arrays are composed predominantly of bedding-parallel veins linked by smaller cross fractures that dip northward at 60° and that represent the smallest end member of a spectrum of similarly oriented fractures, ranging up to the east-west–trending basin-bounding normal faults that form along the southern margin of the Bristol

Channel Basin. The fluid pressures were clearly of sufficient magnitude to open these steeply dipping fractures, as well as those parallel with bedding (see Delaney et al., 1986).

I suggest that this demonstration of the way in which aqueous fluids can move through thick, low-permeability successions is equally applicable to the migration of hydrocarbons through such successions during the evolution of a basin.

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